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Abstract

Despite a long history of glaciological research, the palaeo-environmental significance of moraine systems in the Kebnekaise Mountains, Sweden, has remained uncertain. These landforms offer the potential to elucidate glacier response prior to the period of direct monitoring and provide an insight into the ice-marginal processes operating at polythermal valley glaciers. This study set out to test existing interpretations of Scandinavian ice-marginal moraines, which invoke ice stagnation, pushing, stacking/dumping and push-deformation as important moraine forming processes. Moraines at Isfallsglaciären were investigated using ground-penetrating radar to document the internal structural characteristics of the landform assemblage. Radar surveys revealed a range of substrate composition and reflectors, indicating a debris-ice interface and bounding surfaces within the moraine. The moraine is demonstrated to contain both ice-rich and debris-rich zones, reflecting a complex depositional history and a polygenetic origin. As a consequence of glacier overriding, the morphology of these landforms provides a misleading indicator of glacial history. Traditional geochronological methods are unlikely to be effective on this type of land- form as the fresh surface may post-date the formation of the landform following reoccupation of the moraine rampart by the glacier. This research highlights that the interpretation of geochronological data sets from similar moraine systems should be undertaken with caution.

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Introduction

The moraines developed across Scandinavia have a long history of geomorphological research (e.g. Schytt 1959; Karl,en 1973; Shakesby et al. 1987; Matthews et al. 1995; Etienne et al. 2003; Hayman & Hättestrand 2006; Winkler & Matthews 2010; Matthews et al. 2014), and have served as early study sites for ice-cored landforms (Østrem 1959, 1963, 1964, 1965; Ackert 1984). Specifically, Østrem (1964) recognised that some moraines in Scandinavia are disproportionately large in size compared to the glaciers that formed them, suggesting the presence of buried ice. The formation of such moraine complexes has been subject to uncertainty surrounding: (i) the origin of ice (Schytt 1959; Østrem 1963, 1964; Ackert 1984); (ii) the distinction between ice-marginal moraine complexes and rock glaciers (Barsch 1971; Østrem 1971); and (iii) the interaction between existing moraine ramparts and advancing glaciers in permafrost terrain (Matthews & Shakesby 1984; Shakesby et al. 1987, 2004; Matthews et al. 2014). Ice-cored moraines are commonly found at the margins of terrestrially terminating glaciers (Krüger & Kjær 2000; Schomacker & Kjær 2008; Midgley et al. 2013; Tonkin et al. 2016); however, the long-term preservation of ice is typically limited under an ameliorating climate. In areas characterized by permafrost, ice-cored terrain can be preserved over longer time scales (Sugden et al. 1995). A typical ice-cored moraine is composed of relict glacier ice that becomes isolated from the glacier terminus under a sufficiently thick debris cover (Goldthwait 1951; Østrem 1959, 1964; Evans 2009). Whilst researchers have observed glacier ice within the structure of ice-marginal moraines in Scandinavia (Schytt 1959; Ackert 1984), after analysing the crystallographic properties, Østrem (1963) highlighted that ice within moraines could be of meteoric origin, and may originate as a moraine distal snowbank that was subsequently overridden by an advancing glacier and incorporated into the internal structure of ice- marginal landforms (Østrem 1963, 1964). However, Østrem (1964) also acknowledged that buried ice may have a complex origin, with the potential for stagnating glacier ice to also be incorporated into moraine structure (e.g. 'controlled moraine'; see Evans 2009), but considered that to some extent, most large moraines contained varying quantities of snowbank ice. In recent years, the term 'Østrem' type moraine has been introduced in the literature (e.g. Whalley 2009) to distinguish these moraine systems from ice-cored moraine counterparts in the high-Arctic that predominantly contain glacier ice (Evans 2009). Ice-cored moraines have also been interpreted as rock glaciers (e.g. Barsch 1971). Whilst a polygenetic interpretation could be appropriate for some ice-cored moraines that may transition into rock glaciers (Whalley & Martin 1992; Berthling 2011), geomorphologically stable features deposited on near-level terrain were cited by Østrem (1971) as suitable criteria for classifying features as ice-cored moraine.

Karlen (1973) argued for 'proximal enlargement' as an important moraine forming process in northern Sweden. Karlen envisaged a scenario where moraine ramparts acted as a topographic barrier for subsequent glacier advances, leading to the incremental stacking of imbricate 'drift sheets' onto ice-proximal slopes. These 'drift sheets' were proposed to correspond to successive episodes of Holocene glacier expansion. This hypothesis was favoured despite ground-level photographical evidence from c. 1910 depicting various glaciers in northern Sweden partially overriding their respective moraine complexes (Karlen 1973).

Conversely, in southern Norway a 'push deformation' hypothesis has been proposed in which post depositional modification of existing moraine results from a subsequent glacier advance. The transmission of stress is suggested to result in moraine complexes with a series of anastomosing ridges, and steep proximal and distal slope angles (Shakesby et al. 1987). This hypothesis, which differs from Østrem (1964) who considered the overriding and distal deposition of ridges as an important moraine forming process, has been demonstrated to be important in the Breheimen and Jotunheimen regions of southern Norway (Matthews et al. 2014). However, Matthews et al. (2014) highlighted that additional geophysical survey work is required to validate the exact mechanisms of moraine formation and modification.

The glaciers of the Kebnekaise region have been subject to significant glaciological research (e.g. Schytt 1962, 1966; Holmlund et al. 1996; Holmlund & Jansson 1999; Zemp et al. 2010; Rippin et al. 2011; Gusmeroli et al. 2012; Brugger & Pankratz 2015). However, despite considerable research also investigating moraine development (e.g. Østrem 1964; Karlen 1973; Etienne et al. 2003; Heyman & H€attestrand 2006), the full palaeo-environmental and glaciological significance of the landforms remains unclear. This is at odds with the importance of these sites for contextualizing current and future glacier change. The potential snowbank origin of ice contained within these moraines, the various competing hypotheses in relation to their mode of formation and the potential post depositional modification of these landforms distinguishes them from other ice-cored moraine, yet there is a paucity of research that investigates the significance of these geomorphological features. A modern investigation of the characteristics of these features is therefore warranted.

This study set out to test existing interpretations of Scandinavian ice-marginal moraines, which invoke ice stagnation, pushing, stacking/dumping and deformation as important moraine forming processes. The objectives of this research were therefore to: (i) document the structural character of moraines at Isfallsglaciären using ground-penetrating radar (GPR); (ii) infer the mode of formation and palaeoglaciological significance of the moraine complex developed at Isfallsglaciären; and (iii) examine the wider implications in relation to the use of Scandinavian moraines as a palaeo-environmental proxy. This work is important because Isfallsglaciären has been dynamic over the course of the Holocene and the moraines provide insight into these changes prior to the period of direct measurements and observations.

Overview of study site

Isfallsglaciären is a ~1.5-km long valley glacier located in the Kebnekaise Mountains in northern Sweden (Fig. 1). The glacier has an easterly aspect and has receded ~500 m from the recent maximum extent in the 1920s, when the glacier overrode the inner moraine ridge (e.g. Østrem 1963; Karl,en 1973). Like the neighbouring Storglaciären, Isfallsglaciären is polythermal in character (Eklund & Hart 1996). Schytt (1962), for example, recorded subfreezing temperatures in an artificially created tunnel at the glacier terminus. Storglaciären is currently undergoing changes to its thermal configuration (Pettersson et al. 2003), with one third of its cold surface layer lost over the 1989– 2009 period (Gusmeroli et al. 2012). These changes have been linked to recent climatic amelioration, such as increased winter air temperatures since the 1980s (Pettersson et al. 2003; Gusmeroli et al. 2012). It is likely that the thermal regime of Isfallsglaciären is undergoing a similar evolution.

The morphology of the Isfallsglaciären moraines has previously been described by Schytt (1959) and Karl,en (1973). In this study, the moraine complex is split into three zones based on morphological criteria (Fig. 1). Within the outer-frontal zone (Zi), a subdued moraine ridge attains a relief of ~10 m above the surrounding terrain. A series of discontinuous mounds are present on the distal slope of this ridge. Within the inner-frontal zone (Zii), a moraine ridge rises up to ~20 m above the surrounding terrain. Moraines in Zii are over-printed with flutes related to overriding of the ridge by a glacier advance in 1910 (Karlen 1973). Within the lateral complex (Ziii), a lateral moraine of significant topographical prominence rises ~20–30 m above the surrounding terrain. This feature displays a furrowed morphology and includes a prominent arcuate ridge. A semi-permanent snowbank occurs on the distal slope of this feature (e.g. Østrem 1964; Karlen 1973).

Materials and methods

Radar data were collected using a Pulse EKKO Pro GPR in spring 2013 under winter conditions to ensure frozen ground. Reflection surveys were undertaken with a 100 MHz perpendicular broadside antenna configuration using a 0.25-m step size between traces and a 1-m transmitter/receiver separation distance. A distance of >5 m was kept between the control unit and transmitter/receiver setup to minimize signal interference. Traces were manually triggered using either the control unit interface or a CANBUS electrical beeper and used a time window of 800 ns. Surveys were conducted along a 100-m tape to ensure that the correct step-size was maintained throughout the survey. To correct radar profiles for topography, height was surveyed on each transect using an automatic level. A Garmin GPSMap 62 was also used to record the start and finish location of each transect. During common mid-point/wide-angle reflection-refraction (CMP/ WARR) surveys, the fibre optic cables (each of which were 20 m in length) limited the maximum separation to 38 m, with WARR surveys using a common receiver configuration. Post-processing of CMP/WARR and reflection data was conducted using the EKKO_View Deluxe software from Sensors and Software. Three post-processes were applied to reflection profiles: (i) dewow; (ii) topographical correction; and (iii) gain control. Automatic gain control (AGC) was applied to six of the seven survey profiles. For a single profile, constant gain was found to provide a clearer visualization of subsurface features, so it was used in place of AGC.

The radar data were interpreted qualitatively following post-processing. Terminology used to describe radar facies and surfaces was adopted from Neal (2004), Pellicer & Gibson (2011) and Lindhorst & Schutter (2014). Four main characteristics for reflectors were noted: (i) the reflector shape (planar, wavy, convex, concave); (ii) the reflector dip (horizontal, or either up- or down-glacier dipping); (iii) the relationship between different reflectors within a radargram (parallel, subparallel, oblique, chaotic); and (iv) the continuity of reflectors within a radargram (continuous, moderately continuous or discontinuous). Sedimentology was assessed under summer conditions via shallow excavations (<1 m) to 'ground truth' the observed radar-facies. Facies were assessed using the Hambrey (1994) classification for poorly sorted sediments. Clasts (samples of n = 50 per facies) were assessed for shape and roundness (Powers 1953; Benn 2004); however, it was not feasible to record clast shape for boulder-gravel facies. Shape (C40) and roundness (RA) indices were calculated to facilitate discrimination between samples (e.g. Benn & Ballantyne 1994). These data are presented alongside the reflection data sets.

Results

Radar propagation velocity

WARR and CMP data sets were processed to obtain radar-wave velocities (Fig. 2). Two surveys were completed per zone (Zi, Zii and Ziii; Fig. 1). Outer-frontal (Zi) surveys (A and B) and inner-frontal (Zii) surveys (C and D) all provided similar radar velocities at ~0.11 m ns-1. The velocities were located at a two-way travel time of <150 ns, indicating strong signal attenuation. The lateral complex (Ziii) was broadly found to exhibit higher propagation velocities. Specifically, values of ~0.15 m ns-1 are identified on both lateral complex surveys (E and F). These are detected at a time window of 100–300 ns.

Reflection surveys and surficial sedimentology

Outer-frontal (Zi). – Profiles 1 and 2 both ran transverse to the ridge crestline in the outer-frontal zone (Zi) (distal slope to the right; Fig. 3; Table 1). Profile 1 appeared to be more structurally diverse, with the proximal slope of the landform intersected by a clear up-glacier dipping reflector. Ground truth surveys under summer conditions found a topographically prominent facies of mud on the proximal slope, which related to this radar surface. Below this feature a series of discontinuous, up-glacier dipping reflectors were also visible. Reflectors within the crest of the landform were irregular and hyperbolic, and corresponded to deposits of diamicton (% RA = 52; C40 = 16). On the distal slope of this landform, coherent, continuous reflectors were visible within a topographically prominent hummock (profile 1). Summer surveys found gravel (% RA

= 40; C40 = 10) interspersed with stratified granular and sandy beds. With the application of AGC, structure was poorly defined at depth within the main ridge.

Profile 2 displayed multiple overlapping hyperbolic point diffractions and irregular medium and high amplitude reflectors. Unlike profile 1, a partially coherent down-glacier dipping reflector appeared to dissect the feature between ~27 and 35 m and also corresponded with a change in surface morphology. Excavations revealed that diamicton was present both above and below this reflector. The upper diamicton facies was found to contain a higher percentage of angular clasts (% RA = 76; C40 = 28 and % RA = 72; % C40 = 16) than the lower unit (% RA = 62; C40 = 14 and % RA = 60; C40 = 12). Up-glacier dipping reflectors were also present at depth within the landform and could be seen ~20–30 and ~40–50 m along the profile.

Inner-frontal (Zii). – The inner-frontal ridge was surveyed in profile 3. Shallow excavations (<1 m) along this feature uncovered diamicton with a subangular clast component (% RA = 54; % C40 = 12). The crestline of profile 3 was overprinted with subglacial flutes. The main features of structural interest within this profile were coherent, high amplitude reflectors, which were visible between ~67 and 96 m. The reflectors initially ran subparallel with the moraine surface, before dipping down-glacier. A second less coherent reflector was present at 88–96 m.

Lateral complex (Ziii). – Profile 4 covered the area where the frontal and lateral moraine sections of the landform adjoined. Atypical of other reflection surveys, continuous reflectors could be seen running sub-parallel to the moraine surface. At ~28 m along this transect two coherent reflectors could be seen to cross-cut each other. A radar facies characterized by irregular reflectors and overlapping hyperbolic point diffractions was seen both above and below these coherent reflectors. The ridge crest was found to contain a facies of diamicton with a subangular component (% RA = 46; % C40 = 32).

Profiles 5, 6 and 7 displayed the subsurface structure of the southern-lateral complex. Profile 5 ran oblique to the landform (but approximately parallel to the inferred direction of former ice flow). The sedimentology along profiles 5–7 was predominantly angular boulder-gravel with the exception of the small ridge captured in profile 5, which contained diamicton (% RA = 62; C40 = 14) and gravel (% RA = 66; C40 = 16) with down-glacier dipping granular lenses. Similar to other profiles, profile 5 displayed hyperbolic (related to subsurface point diffractions) and irregular reflectors. Two coherent sub-surface reflectors initially ran approximately parallel to the moraine surface, but subsequently dipped down-glacier. These were visible between ~0–14 and 25–47 m along profile 5, respectively.

Profile 6 ran transverse to the southern lateral moraine complex. Here, the main structural feature was a moderately continuous reflector at depth within the moraine. This reflector appeared to run subparallel to the moraine surface and was both over- and underlain by hyperbolic, chaotic and irregular radar facies. A rounded response occurred mid profile (~45–55 m). A snowbank could be distinguished on the ice-distal slope of the landform. The base of the ice-distal slope was characterized by multiple strong point diffractions. Profile 7 ran approximately parallel to the ridge crest of the southern-lateral complex. The main structural feature of interest within this profile could be seen between ~50 and ~130 m along the profile and was located in the ~50 to 70 ns time window. This feature ran subparallel to the moraine surface and appeared to dissect an upper radar-facies consisting of hyperbolic and irregular point diffractions. A less coherent (partially due to the hyperbolic nature of the shallower radar-facies) continuation of this radar surface was present 0 to 20 m along profile 7 at ~45 ns. The near-surface sedimentology along profiles 6 and 7 was predominantly boulder-gravel and diamicton. The percentage of subangular clasts within the boulder-gravel surface facies across Zii increased progressively down-moraine (% RA = 96, 98, 88, 66 and 48). Diamicton sampled from facies in proximity to profiles 6 and 7 had a predominantly angular clast component (% RA = 74; C40 = 20 and 74; C40 = 26).

Interpretation

Radar wave velocities and likely composition

The propagation velocity of radar waves is related to the subsurface composition (Neal 2004). Thus, by relating the velocities obtained to values for known substrates (Table 2), the sedimentological characteristics of the landforms can be inferred. Radar-wave propagation velocity also varies depending on the saturation and thermal state (e.g. frozen or unfrozen) of a material (Neal 2004). Here, moraine composition appears to vary spatially across the lateral-frontal complex. The surveys undertaken in the inner-frontal (Zii) and outer-frontal (Zi) zones indicate that these areas are debris-rich. Schwamborn et al. (2008) found frozen diamicton (with 10% pore water) to have a radar-wave velocity of 0.125 m ns-1 (determined from a CMP survey). This contrasts with unfrozen diamictons and till, which exhibit propagation velocities of 0.06–0.09 m ns-1 (e.g. Burki et al. 2009; Lukas & Sass 2011). Given that the moraines were frozen at the time of the survey, slightly higher velocities are to be expected, especially if sediment is partially saturated prior to winter freezing. Velocities recorded from the inner-frontal (Zii) and outer-frontal (Zi) zones (surveys A–D; ~0.11 m ns-1) are, therefore, consistent with a compo- sition of diamicton with a limited volume of interstitial ice; a finding also consistent with surveys of surface sedimentology (Table 1). It is unclear whether the strong signal attenuation resulting from the thick silt-rich diamicton facies (hence shallow coherent reflections visible in the velocity-depth plots) is masking ice-rich permafrost at depth within the topographically prominent inner-frontal moraine (Zii).

The structural composition of the lateral complex (Ziii) is less straightforward, but is highly likely to indicate the presence of ice within the landform. Here, the wide range of radar propagation velocities (Fig. 2) most likely results from variability in the porosity, amount of interstitial ice and fine material within the landform. Østrem (1963, 1964) directly observed ice within the southern lateral moraine by excavating a series of pits. More recently, Kneisel (2010) detected ice-rich permafrost in moraine at Isfallsglaciären using electrical resistivity tomography (ERT) with surveys undertaken where the lateral and frontal moraines adjoin (C Kneisel, pers. comm. 2013). Given the coarse nature of the surficial sediments (boulder-gravel facies) and known inclusion of ice within the landform, radar-wave velocities derived from rock glaciers (Table 2) are likely to serve as a useful proxy for subsurface composition. For example, Monnier & Kinnard (2013) regarded velocities of 0.15–0.17 m ns-1 within surficial deposits of rock glaciers as evidence of significant quantities of air (high porosity), and calculated that a velocity of 0.16 m ns-1 was equivalent to 22% air content. High porosity may explain high velocities near the surface of the lateral complex (Ziii) given the surface sedimentology of boulder-gravel; however, similar velocities are also identified at depth (a time window in excess of 100 ns) within the landform. Buried ice at the margins of high-Arctic glaciers also results in velocities of 0.15–0.17 m ns-1 (Brandt et al. 2007; Midgley et al. 2013). However, the ice within the lateral zone (Ziii) may have a complex origin, and contain both glacier ice and moraine distal snowbank ice (Østrem 1964). Surveys indicate contrasting velocities to that expected in snow (e.g. Table 2). As such, if snow was included into the structure of the landform, it is likely to be of considerable age (potential age ranging from centuries to millennia; e.g. Karlen 1973), resulting in recrystallization, compression and mixing with debris, thus accounting for lower than expected radar-wave propagation velocities for snow as specified in Table 2. An ice-rich substrate in lateral zones would also be consistent with existing interpretations of the adjacent proglacial zone of Storglaciären, where there is disparity in size between the subdued boulder-rich frontal moraine, and larger lateral landforms (Østrem 1964; Karl,en 1973; Ackert 1984; Etienne et al. 2003). Interestingly, at Storglaciären, the true-right lateral moraine is noted to have undergone postdepositional modification and slope movement (Karl,en 1973; Etienne et al. 2003), which is indicative of an ice-rich substrate, with ice facilitating the transition from moraine to rock glacier.

Internal structure and sedimentology

Outer-frontal (Zi). – Profile 1 provides a clear example of sedimentary units deposited on the ice-proximal slope of an existing moraine ridge. Here, the main moraine ridge contained diamicton as demonstrated by radar propagation velocity surveys and direct observation. Subsequent recession of the glacier margin is suggested to have formed a terrace of massive mud within a low energy depositional environment (e.g. an ice-marginal lake). This sedimentary unit was documented in the field, and also appears as a distinct structural unit in profile 1 (‘mud’; see Figs 3, 4). The moraine hummock on the distal slope exhibited subhorizontal reflectors, which were found to relate to facies of stratified gravel and sand during ground truth surveys. This unit is interpreted as an ice-contact fan resulting from both gravitational flows and glacifluvial deposition; however, the relative chronology in relation to other sections of the moraine is unclear (Fig. 4) without further excavation. Ice-proximal deposition appears to be spatially limited across the ridge. The morphological and structural relationships between these sedimentary units suggest that at profile 2 the glacier partially overrode an existing ridge, resulting in two stacked units of diamicton of different ages with a surface contact visible both in-the-field and within the reflection profile.

Inner-frontal (Zii). – High concentrations of silt – such as are present in many diamictons – are associ- ated with poor signal penetration (e.g. Overgaard & Jakobsen 2001). Given the presence of diamicton with the moraines, the strong levels of signal attenuation at depth within the inner-frontal zone (Zii) is highly likely to indicate high silt content within the matrix of the diamicton; a finding consistent with field surveys. The geometry of the radar surfaces (down-glacier dipping) documented here are not consistent with the conceptual model produced by Karlen (1973) who suggested that structurally, moraines largely consist of imbricately arranged units of poorly sorted glacial sediment (‘drift sheets’). The origin of the down-glacier dipping structures are uncertain, but assuming that the frontal moraine is ice free (e.g. Østrem 1964; the CMP/WARR data presented here and the high levels of attenuation seen in the reflection survey), the surfaces may relate to bounding layers between stacked units of diamicton. This interpretation would require multiple periods of moraine development and partial overriding of existing moraine obstacles, rather than the proximal enlargement model envisaged by Karlen (1973). Evidence such as the higher levels of clast subangularity in this zone (indicative of subglacial processes), the large size of the frontal moraine, the overprinting of flutes and evidence of overriding of the frontal moraine in 1910, highlight that this landform has a complex origin resulting from multiple periods of development.

Lateral complex (Ziii). – For Ziii, the hyperbolic radar facies seen in this zone are interpreted as evidence of a predominantly coarse and massive structural configuration, which is consistent with coarse deposits of boulder-gravel found on the moraine surface. Superimposed ridges (e.g. as seen in profile 5) and similar dipping structures to those documented on the frontal-ridge are interpreted as evidence of overriding and distal deposition of material by the glacier on the southern-lateral complex in a similar manner to that envisaged in Zii. Small moraine ridges such as the arcuate ridge visible in profile 5 could have developed in response to the dumping, pushing or squeezing of material at the ice margin (e.g. Price 1970; Birnie 1977; Boulton & Eyles 1979; Bennett 2001; Krüger et al. 2010), or the freeze-on of sediment related to annual oscillations of the ice front (Krüger 1995). Pushing as a moraine forming mechanism is unlikely here as: (i) dominant ice-proximal sediments are dissimilar to those contained within the ridge; (ii) coarse boulder facies have high shear strengths and thus are not particularly conducive to push moraine formation (Cook et al. 2013); and (iii) the ridge contains diamicton with granular lenses, which are linear in form and lack displacement structures associated with ice-marginal stress.

Here, subhorizontal and rounded reflectors such as those seen in profiles 6 and 7 are likely to indicate the interface between the surficial deposits of diamicton and boulder-gravel, and an ice-rich substrate at depth. Moorman et al. (2003), for example, found that strong continuous reflectors in GPR surveys undertaken in permafrost terrain were related to the interface between frozen and unfrozen ground conditions. An interpretation of ice-rich permafrost is also consistent with the field observations of Østrem (1964), who excavated the southern-lateral complex, and found ice at depths of 2.2, 2.5 and 2.8 m. Here, for example, the estimated depth to the reflector, thus thickness of the upper surface layer in question, ranges between ~2.25 and ~4.5 m in profile 7.

Discussion

Development and significance of the Isfallsglaciären moraines

Conceptually, the moraine system at Isfallsglaciären is clearly distinguishable from alpine temperate glacial landsystems, where distinct asymmetrical ice-contact ramps are produced as a result of the flowage of debris from supraglacial positions (Humlum 1978; Boulton & Eyles 1979; Röthlisberger & Schneebeli 1979; Small 1983; Lukas & Sass 2011; Lukas et al. 2012). The mor- phological characteristics of the moraines share some similarity with multi-crested ‘controlled’ ice-cored moraine complexes documented to occur in some high-Arctic and Icelandic glacial landsystems (Evans 2009, 2010; Ewertowski et al. 2012); however, when compared to the geophysical data sets presented in Midgley et al. (2013) there are distinct differences. Midgley et al. (2013) documented very coherent up-glacier dipping reflectors within ice-cored moraine in the Norwegian high-Arctic, which were interpreted as debris-bearing features contained within buried glacier ice; a stark difference to the hyperbolic structures found here, despite the presence of ice within the lateral complex (Ziii). This discrepancy in moraine structure may lend support for a unique mode of ice-incorporation operating at the margins of polythermal glaciers in northern Sweden (e.g. Østrem 1963, 1964).

The clast-form data set shows a marked compositional decrease in clast angularity from lateral to frontal zones of the moraine system. Clast-form gradients have been recorded at a number of sites in Scandinavia and Iceland where roundness has been found to decrease with distance from the former glacier terminus (Matthews & Petch 1982; Benn & Ballantyne 1994; Spedding & Evans 2002). The clast composition of ice-marginal moraines can relate to the relative importance of passive and active debris transport pathways (e.g. Matthews & Petch 1982; Evans 2010), and the pushing of pre-existing valley side paraglacial debris (e.g. Matthews & Petch 1982). In other areas, the recycling of pre-existing debris by cycles of glacier activity may result in an increase in clast-form 'maturity' (e.g. Burki 2009). In ground-level photography taken by Enqvist in 1910 the glacier surface appears to be relatively free of supraglacial debris leading to well-exposed subglacial sediments within the forefield. Debris can, however, be seen emerging from the ice front, indicating the relative importance of subglacial debris pathways in the frontal zone of the former terminus. The higher proportion of subangular clasts in frontal zones may also demonstrate the importance of processes such as: (i) the accretion of subglacial till onto existing moraine; (ii) the thickening of debris-rich basal ice at the terminus in response to the reverse moraine slope and the cold-based conditions (e.g. Pomeroy 2013); or (iii) the recycling of existing sediment within the glacier forefield (e.g. Burki 2009). Similarities between clast-roundness in profile 3 and profile 4 (e.g. % RA = 46–54), both of which cross-cut more frontal sections of the moraine complex, and control samples from the fluted glacier forefield (% RA = 38, 42 and 44), are likely to reflect the importance of one or more of the above processes.

Whilst Karlen (1973) disregarded the ground-level photography taken by Enquist, arguing that proximal enlargement was an important mechanism of moraine development, the structural characteristics (down-glacier dipping reflectors) lend support to the hypothesis of overtopping and distal deposition of debris (profile 2 in Zi and profile 3 in Zii; see Fig. 4). Assuming that the ice margin remains stable over multiple years, mixtures of debris and snow present on the ice-distal face of the moraine will be incorporated into the structure of the landform (e.g. Østrem 1964). Given the limited supraglacial debris visible in the 1910 ground-level photography, the ice margin would need to remain stationary over a considerable period of time to facilitate moraine construction (Boulton & Eyles 1979; Benn et al. 2003). One issue is that debris run out over distal snowbanks is only observed on frontal sections of the moraine in the 1910 ground-level photography (locations approximately delimited in Fig. 4). However, based on interpretation of radar velocity estimates, ice is principally located in the lateral zone (Ziii) of the assessed moraine. The spatial distribution of buried ice inferred from the geophysical surveys presented, however, accord with existing studies in Scandinavia (e.g. Østrem 1964), and also observations in the high-Arctic (e.g. Midgley et al. 2013; Tonkin et al. 2016) where high quantities of buried ice in moraine systems appear to be principally located in lateral ice-marginal areas.

An issue requiring further comment is the topographic influence of a pre-existing moraine on glacier geometry (e.g. Spedding & Evans 2002; Barr & Lovell 2014). For the neighbouring Storglaciären, initial moraine formation c. 2.5 ka BP is suggested (Karlen 1973; Ackert 1984; Etienne et al. 2003). On the assumption that the adjacent Isfallsglaciären moraines formed simultaneously, the landforms are highly likely to have exerted a topographic influence on later glacier advance stages. A range of recent Holocene Neoglacial advances between 2.7–2.0, 1.9–1.6, 1.2–1.0 and 0.7–0.2 ka BP were suggested for Scandinavian glaciers by Karlen & Kuylenstierna (1996) with valley glaciers attaining their largest Neoglacial extent during the 17th and 18th centuries (e.g. Karlen 1988; Nesje 2009). Ice-marginal positions demarcated by the previously discussed historical ground-level photography (e.g. photographs taken by Enqvist in 1910; see Fig. 4) and by measurements from 1915 provided by Hamberg et al. (1930) highlight sustained overriding of the inner moraine ridge over a 5-year period between 1910 and 1915 (see Schytt 1959 for a review of historical glacier records). The historical imagery, therefore, demonstrates that overriding is important for the development of the inner-ridge, which, if considered alongside the bounding surfaces identified in radar profile 3, may have occurred at several points in time, resulting in a composite ridge overprinted with flutes (profile 3 in Fig. 4). A discrepancy between moraine size and debris production rate has been suggested to indicate landform development over a time scale in excess of the Little Ice Age at other sites in Scandinavia (Matthews & Petch 1982). Given that the moraines at Isfallsglaciären are likely to also have been developed over long time scales, the push-deformation model as envisaged for a number of high-alpine moraine systems in southern Norway (e.g. Matthews & Shakesby 1984; Shakesby et al. 1987, 2004; Matthews et al. 2014) may also be relevant to Isfallsglaciären, although it is at odds with a model of overriding, which is clearly an important geomorphological process for certain sections (e.g. Zi and Zii) of the Isfallsglaciären moraine system. In summary, the evidence presented in this research highlights that the moraines are likely to be polygenetic in origin, as indicated by observed differences in the internal character and sedimentology across the moraine complex, and time transgressive in age. The resulting geomorphology is a product of the repeated reoccupation of the moraine system by glacier ice, and thus has resulted in what can be described as a ‘palimpsest’ landsystem (e.g. Pomeroy 2013). The morphological and subsurface characteristics of the outer-frontal moraine (Zi) especially illustrate the composite character of the surveyed moraines.

Reconciling existing geochronologies and structural data

The geochronology of the moraines is currently poorly constrained. The reliability of lichenometric dates from the Isfallsglaciären forefield (e.g. Karlen 1973) are potentially unclear as: (i) the moraine system has been both partially and fully overridden, resulting in the reworking of surface materials; and (ii) the moraine appears to have been subject to extensive snow cover, which at other sites has been linked to reduced confidence in the reliability of lichenometric ages (e.g. Benedict 1993; Osborn et al. 2015). Hormes et al. (2004) presented radiocarbon dates from a small valley glacier ~6 km north of Isfallsglaciären. Unlike many moraine systems in the Kebnekaise region, palaeosols were identified within the stratigraphy of these landforms. From the analysis of organic material, Hormes et al. (2004) advocated four periods of soil formation at Nipalsglaciären: 7.8–7.58, 6.3–4.08, 2.45–2.0 and 1.17–0.74 cal. ka BP. Broadly similar responses of Isfallsglaciären to climatic variability during these periods are likely, although it is acknowledged that the two glaciers may have responded differently to environmental change due to site-specific aspect, hypsometry and topo-climate. In the absence of robust dating controls at Isfallsglaciären, and issues with existing lichenometric dates due to the processes of partial glacier self-censoring (e.g. Gibbons et al. 1984; Kirkbride & Winkler 2012), over-extrapolation and snow cover (Osborn et al. 2015), moraine chronologies remain uncertain. Further work could apply additional dating controls; however, it is argued that issues related to the recycling of glacigenic debris (e.g. Burki 2009) resulting from the overriding of pre-existing materials and potential glacier-permafrost interactions (e.g. Etzelümller & Hagen 2005; Matthews et al. 2014) are likely to result in problematic or inconclusive data sets. On this type of landform, traditional geochronological techniques are unsuitable, as the surface appears to post-date the landform (e.g. as indicated by the recent overprinting by flutes). Based on interpretation of the structural data reported here, morphology alone is a poor indicator of glacial history due to the palimpsest nature of the landforms (e.g. Fig. 4), which, despite a probable formation during the mid-late

Holocene, have survived periods of glacier re-advance, indicating ineffective self-censoring. As such, great care needs to be employed when interpreting geochronological data obtained from similar landforms. Additional work to document the structural characteristics of a wider range of Scandinavian moraines is a worthwhile endeavour, which may further advance understanding of the glaciological significance of the moraines and facilitate understanding of relict landform assemblages in the geomorphological record.

Conclusions

Radar propagation velocity surveys reported here highlight that the frontal zone of the moraine system at Isfallsglaciären is debris-rich, whereas lateral zones are ice-rich. The lateral zones of the moraines are, however, structurally divergent from ice-cored moraine counterparts in the high-Arctic, revealing hyperbolic and chaotic radar facies. GPR reflection profiles appear to demarcate the spatial extent and depth at which ice within the southern-lateral complex (Ziii) is buried. Radar-depth conversions are in broad agreement with the reported findings of Østrem (1964), indicating ice at depths of ~2.25–4.50 m along profile 7. Given that previously destructive methods were used to investigate moraine structure, GPR is shown to be a valuable tool for documenting the structure of ice-marginal landforms. The frontal landforms are complex in form, and whilst in places (e.g. parts of Zi) the ridges may represent former ice-marginal positions resulting from Holocene glacier re-advances, in others, overriding has reworked surface materials (e.g. Zii). It is unclear whether the transmission of stress onto pre-existing ridges has influenced the morphology of the investigated moraine system. The application of traditional geochronological methods are unlikely to provide a useful measure of moraine age due to the palimpsest nature of the moraine system, which has most likely developed as a result of repeated occupation of the glacier forefield over the course of the Holocene. It is hoped that the data set provided here will not only facilitate greater understanding of the geochronology, and likely mode of formation of ice-marginal moraine at polythermal glaciers, but also aid interpretations of relict landform assemblages in glaciated valley landsystems.

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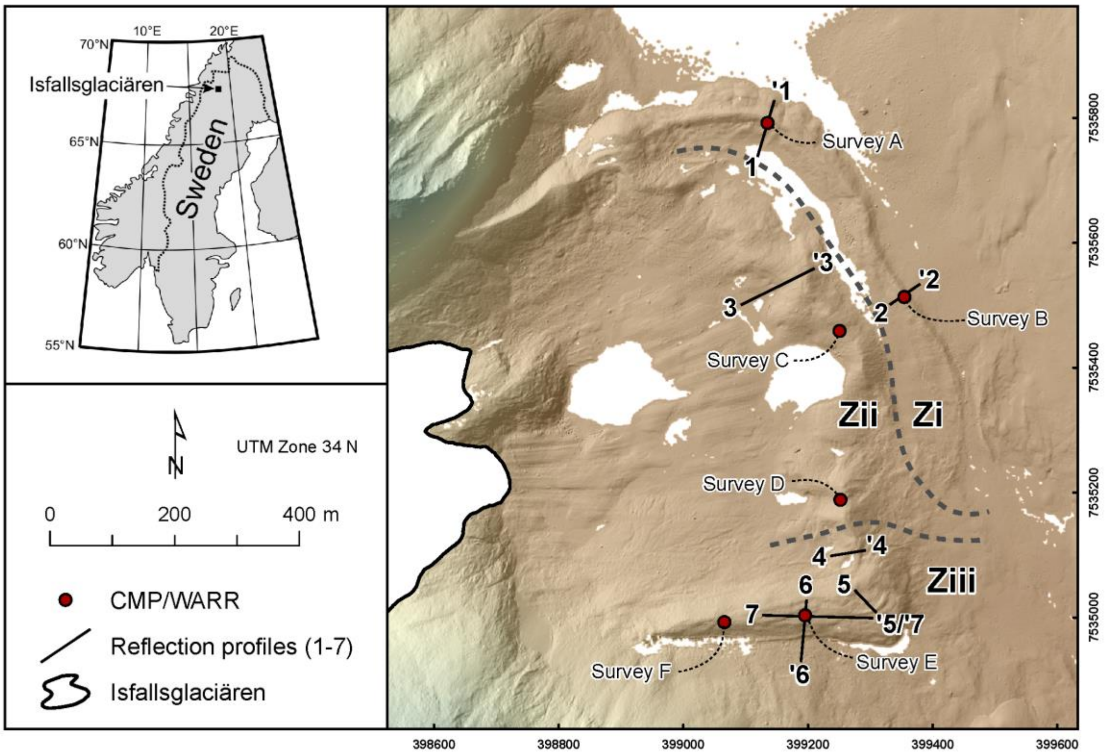
Fig. 1. The location of Isfallsglaciären in relation to Scandinavia. The locations of the geophysical surveys reported are displayed over a hillshaded model of the moraines produced from data provided by Carrivick et al. (2015).

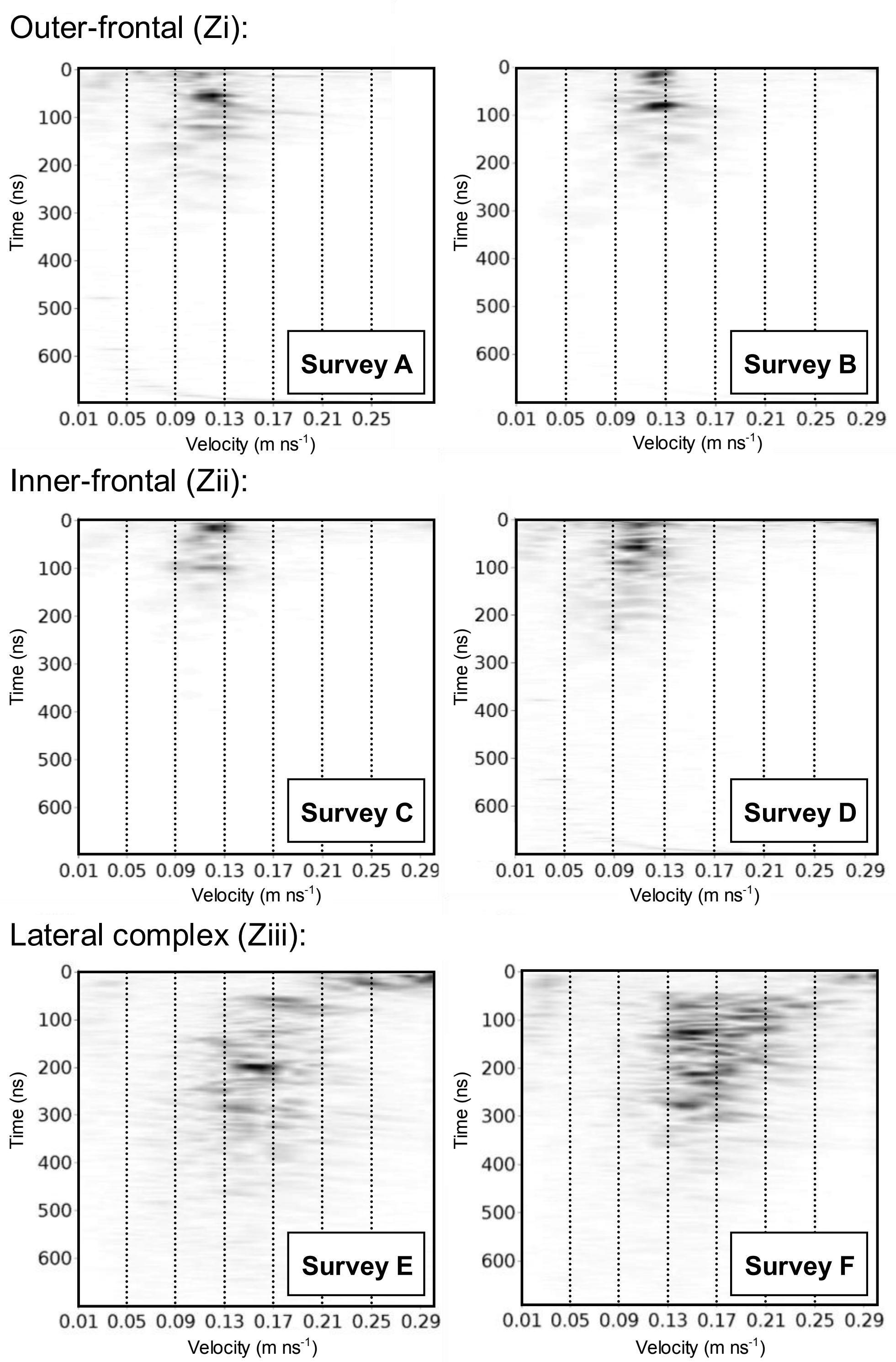
Fig. 2. Plots showing the radar wave propagation velocity at different locations across the moraine complex. Note how propagation velocity is increased at lateral positions.

Fig. 3. Topographically migrated GPR reflection surveys. Profiles 1–6 cross cut the moraine crestline, with the ice-proximal slope to the right. Approximately 25 m along profile 6, the transect cuts across profile 7. Profile 7 runs approximately parallel to the moraine crestline with up-glacier sections of the landform to the right, and crosses profile 6 at ~68 m along the transect. Profile 7 has not been adjusted for topography.

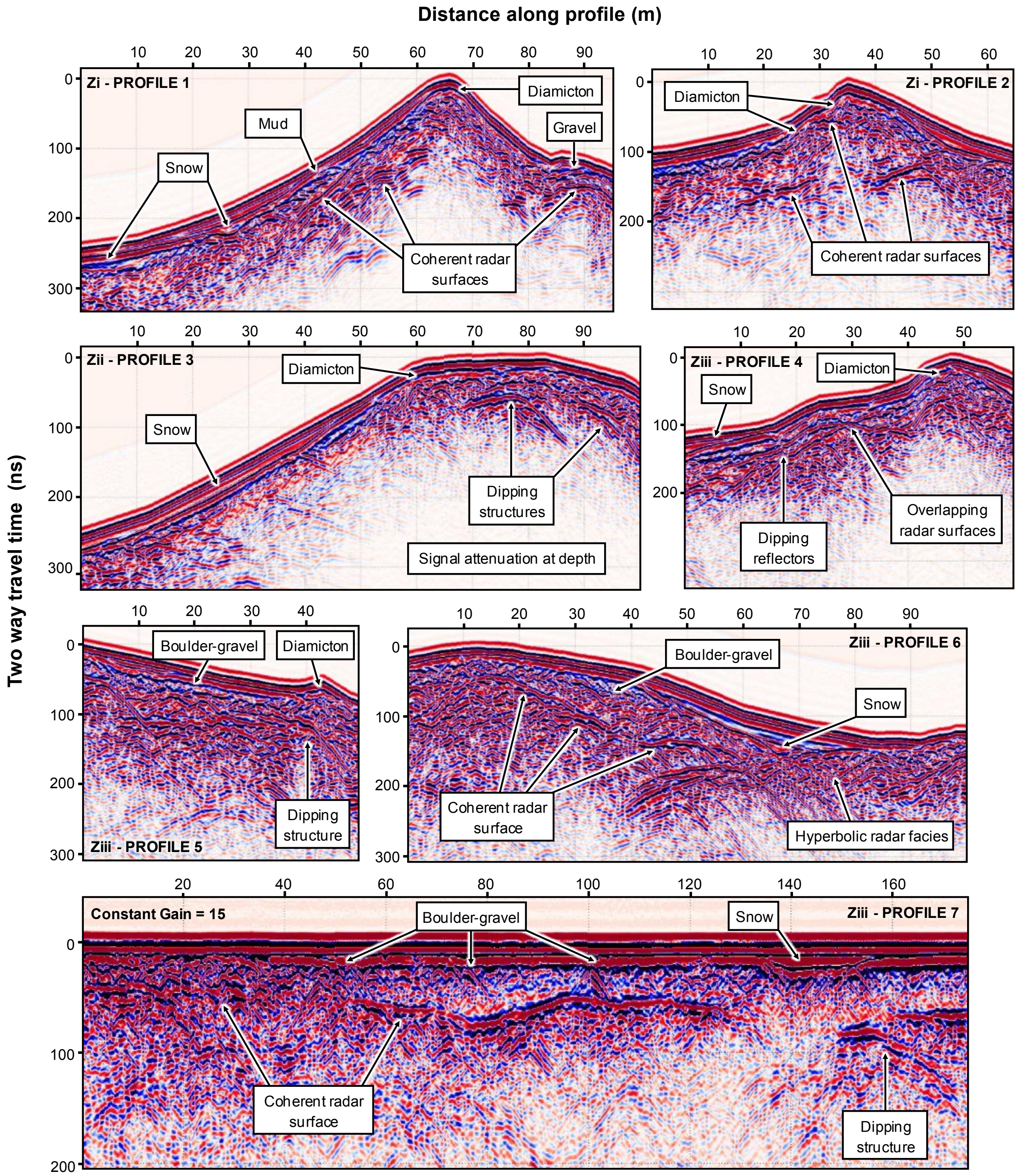
Fig. 4. An illustration of the former glacier in 1910 and an interpretation of the ice-marginal moraines within the different zones.



70°N
10°E
20°E
Isfallsglaciären
65°N
Sweden
60°N
55°N
UTM Zone 34 N
0
200
400 m
CMP/WARR
Reflection profiles (1-7)
Isfallsglaciären
'1
1
Survey A
'3
3
'2
2
Survey B
Survey C
Zii
Zi
Survey D
4
'4
Ziii
6
5
7
'5/'7
Survey F
'6
Survey E
7535800
7535600
7535400
7535200
7535000
398600
398800
399000
399200
399400
399600

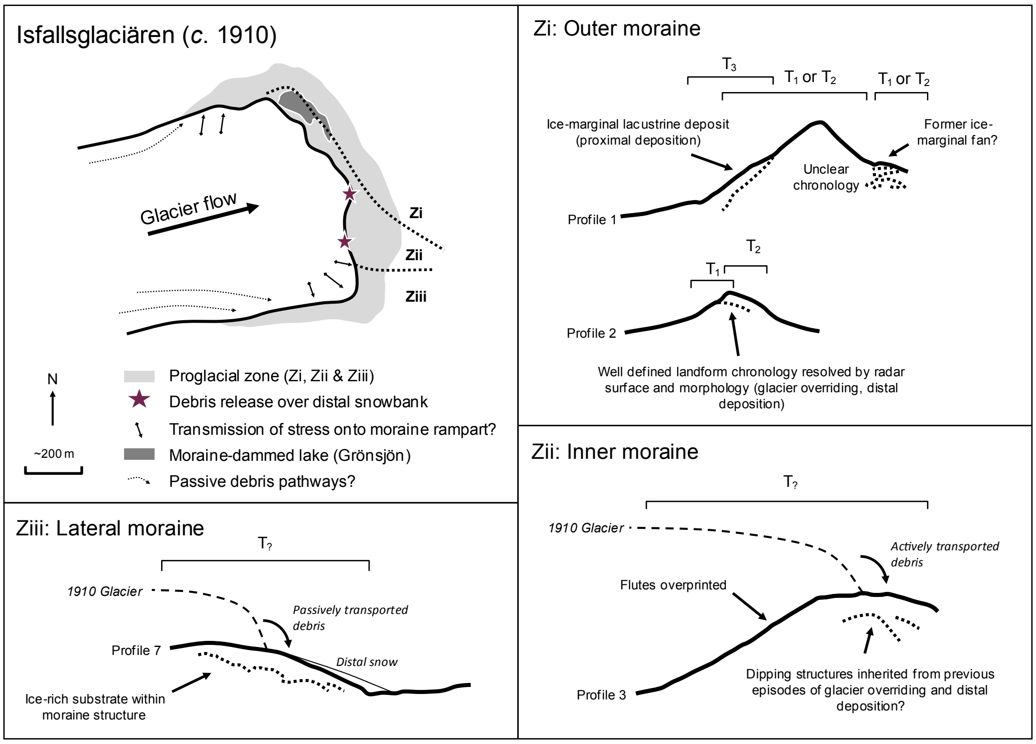


Outer-frontal (Zi):
Survey A
Survey B
Inner-frontal (Zii):
Survey C
Survey D
Lateral complex (Ziii):
Survey E
Survey F
Time (ns)
Velocity (m ns-1)
0
100
200
300
400
500
600
0.01 0.05 0.09 0.13 0.17 0.21 0.25 0.29



Time/depth conversion: Depth = v x TWTT / 2 100 ns = 5 metres (if v = 0.10 m ns^{-1})

AGC applied to all profiles (unless stated) 7.5 metres (if v = 0.15 m ns^{-1})



Isfallsglaciären (c. 1910)
Glacier flow
Zi
Zii
Ziii
N
~200 m
Proglacial zone (Zi, Zii & Ziii)
Debris release over distal snowbank
Transmission of stress onto moraine rampart?
Moraine-dammed lake (Grönsjön)
Passive debris pathways?
Zi: Outer moraine
T3
T1 or T2
T1 or T2
Ice-marginal lacustrine deposit (proximal deposition)
Former ice-marginal fan?
Unclear chronology
Profile 1
T2
T1
Profile 2
Well defined landform chronology resolved by radar surface and morphology (glacier overriding, distal deposition)
Zii: Inner moraine
T?
1910 Glacier
Actively transported debris
Flutes overprinted
Profile 3
Dipping structures inherited from previous episodes of glacier overriding and distal deposition?
Ziii: Lateral moraine
T?
1910 Glacier
Passively transported debris
Profile 7
Distal snow
Ice-rich substrate within moraine structure

Table 1. Summary of interpreted radar facies.

Zone	***Profile***	***Radar-surface geometries***	***Radar facies***	***Signal attenuation***	***Surficial sedimentology***	***Likely composition***
Zi	1	Dipping up glacier; Sub-horizontal to the moraine surface	Chaotic	High	Mud, diamicton, gravel	Debris
	2	Dipping up glacier; Dipping down glacier	Chaotic	High	Diamicton	Debris
Zii	3	Dipping down glacier; Sub-parallel to the moraine surface	Chaotic	High	Diamicton	Debris
Ziii	4	Dipping up glacier; Sub-parallel to the moraine surface	Chaotic	Moderate	Diamicton	Debris-ice mix?
	5	Dipping down glacier; Sub-parallel to the moraine surface	Chaotic; Hyperbolic	Low	Boulder-gravel, diamicton	Debris-ice mix
	6	Dipping down glacier; Sub-parallel to the moraine surface	Chaotic; Hyperbolic	Low	Boulder-gravel, diamicton	Debris-ice mix
	7	Dipping down glacier; Sub-parallel to the moraine surface	Chaotic; Hyperbolic	Low	Boulder-gravel, diamicton	Debris-ice mix

Table 2. Radar velocities for a range of landforms and features.

Substrate	***Velocity (m ns^{-1})***	***Source(s)***
Air	0.3	Reynolds (2011)
Snow	0.194–0.252	Reynolds (2011)
Glacial sediment	0.06–0.10	Sass & Krautblatter (2007); Lukas & Sass (2011); Burki *et al*. (2009)
Diamicton (frozen)	0.115–0.135	Brandt *et al*. (2007); Schwamborn *et al*. (2008)
Loose talus	0.11–0.14	Sass & Krautblatter (2007)
Rock glacier	0.12–0.17	Degenhardt Jr. & Giardino (2003); Degenhardt Jr. (2009); Monnier & Kinnard (2013)
Glacier ice	0.167–0.170	Murray *et al*. (2000)
Buried ice	0.15–0.17	Brandt *et al*. (2007); Midgley *et al*. (2013)